CHAPTER 7

SOIL ORGANIC MATTER AND BIOLOGY IN RELATION TO CLIMATE CHANCE

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ABSTRACT

Climate change, a.o. due to high atmospheric CO_2 concentrations, is expected to produce higher temperatures, and largely unpredictable changes in rainfall. The temperature changes are likely to be least in the tropical zones and largest at high latitudes. Soil carbon changes are implicated both as part of the global carbon cycle, and as a main area of impact of the climatic changes.

Increased CO_2 may lead to more rapid plant growth when this is the currently limiting factor, but it seems unlikely to be so to an important extent except under intensive agriculture. Primary productivity and hence the input of plant residues may then become larger. In addition, higher concentrations of CO_2 may change the quality of residues, and the water use efficiency of vegetation. The CO_2 increase is unlikely to have any significant direct effect within soils, where the CO_2 level is usually enhanced already.

Primary productivity will respond to temperature change only where the latter is limiting, which is least likely in the tropics, but will have a major effect at high latitudes. Rainfall is much more likely to be important in the tropics. The most pervasive effects on soils will be where the whole character of the vegetation changes.

Increased temperature will usually lead to more rapid breakdown of soil organic matter, and these processes have been modeled with promising results. The relatively small changes expected in the tropics appear unlikely to induce key changes, but the effects in high latitudes are expected to be much larger. Soils with high organic matter content due to low temperature and waterlogging could be strongly affected due to both temperature and hydrology changes. The combined effect of changes in organic matter input and in rates of organic material breakdown will set the final new equilibrium level of soil organic matter.

The effects that are most difficult to predict are those resulting from the effects of rainfall changes, and the behavior of soils which are not in equilibrium with the climate. Changes in ecosystems and migration of vegetation zones are likely in some areas, and the speed with which this happens will be of critical importance. Soil flora and fauna, including symbiotic organisms, may be seriously affected by any such changes, as their migration rates are likely to be small.

INTRODUCTION

The general outline of processes leading to climate change is now well known. However the climate predictions from the Global Circulation Models (GCMs) are still uncertain; in particular there are serious doubts about the effects

of cloudiness, and the way in which land-atmosphere interactions should be included in the modeling. There is a reasonable agreement about global temperature changes, but the impacts of climate change will be felt more locally, and the current predictions of regional or national climate change are extremely uncertain. We are, therefore, still very much in the situation of working out the consequences of scenarios that may or may not happen.

Here we start from the most recent predictions of the GCMs (Fig. 7.1) (Dickinson 1986; Trabalka 1985), which predicts temperature increases averaging 1.5-4.5°C. Such changes are likely to be accompanied by major alterations in winds, storms and rainfall. Despite the fact that these proceedings focus most on tropical areas, we cannot ignore the fact that all predictions agree that the temperature effects will be largest at the poles and least at the equator. The higher latitudes must therefore also be considered.

We take as a reasonable range for a possible scenario that rainfall may change by about 20% at any point, though some changes may well be larger than this (Department of the Environment 1988). A point of general agreement is that the increase in temperature will be most marked in winter, so that cold winters, sea, ice, and snow cover are all likely to diminish.

SOIL AS PART OF THE CARBON CYCLE

Soil organic matter occupies a particularly critical position in this topic. The carbon cycle is affected by climate change, and it has an important impact itself upon the changes via the CO_2 in the atmosphere.

The carbon in soil organic matter is a significant component of the earth's carbon reservoirs, with around 1.5×10^{18} g C (Solomon et al. 1985) (Table 7.1). This is small compared to the carbon stored in the deep ocean (ca. 38×10^{18} g C and in the deep earth, but most of this is only in extremely slow equilibrium with the atmosphere, and the amount in the more rapidly equilibrating surface ocean layer is only some 0.6×10^{18} g C. The turnover time of carbon in the soil-vegetation system is variable, ranging from less than a year for annual vegetation, to thousands of years for organic matter in the deeper parts of soil profiles.

The loss of carbon from both terrestrial biomass and soil organic matter has been considerable over the last 200 years. The calculation of the net release, over this whole period or currently per year, is very difficult due to the heterogeneity of the land surface, the variation in treatment or cover, and the uncertainty of the areas of particular ecosystems.

The current net release of C from soils and plants is estimated to be around 10^{15} g C yr^{-1}, but the uncertainties in the data and the variable assumptions and definitions used by different authors provide a very large margin of error.

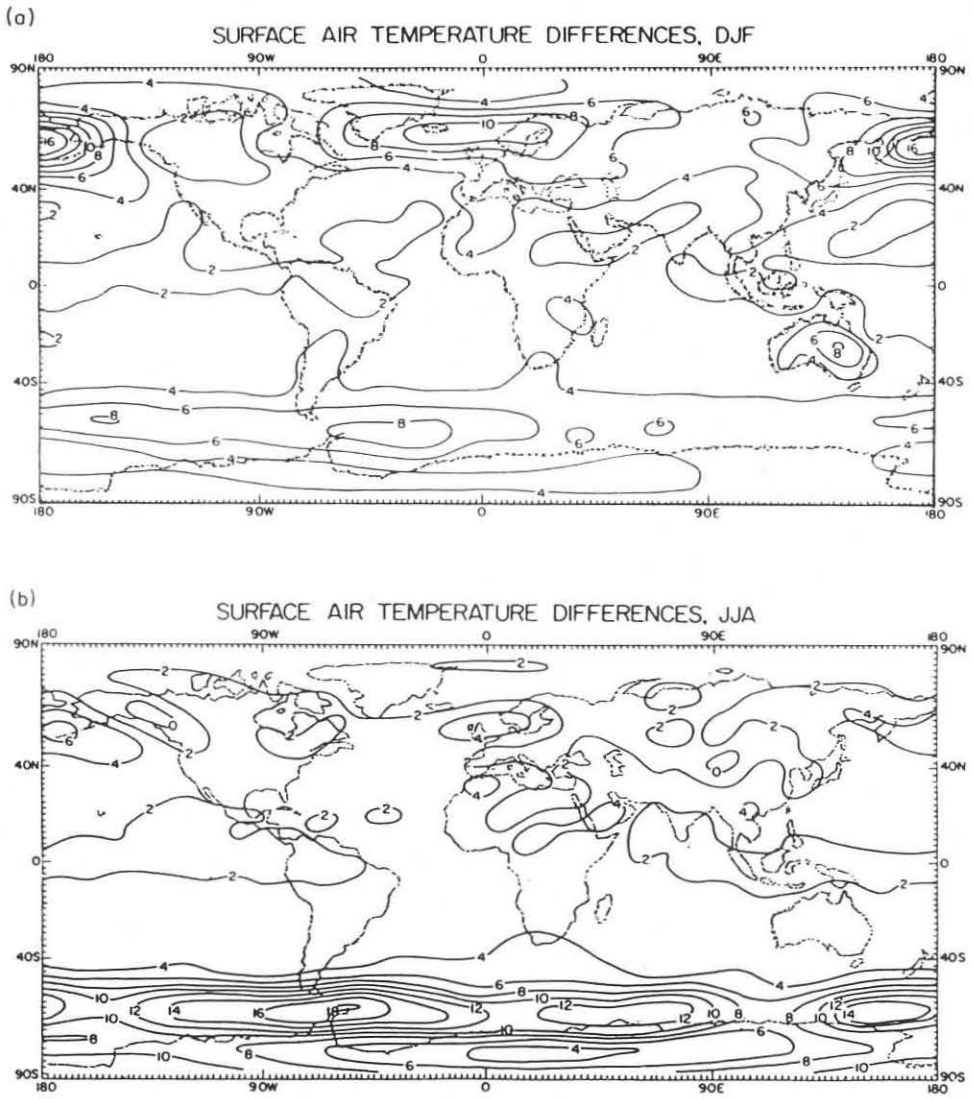

Fig. 7.1 *Predicted air temperature changes (°C) for a steady-state doubling of CO_2, showing winter (a) and summer (b) surface air temperature (from Bolin et al. 1986)*

Estimates of the net annual flux between the atmosphere and terrestrial ecosystems (Houghton et al. 1985) range from -2.0 to 20.0 x 10^{15} g C yr^{-1}, a range that clearly indicates the gross uncertainties that persist in this area. The net release is about 20% of the annual flux of CO_2 from fossil fuel burning, or ca 5 x 10^{15} g C yr^{-1}. At present it is estimated that some 40% of the total increase in

atmospheric CO_2 to date has arisen from changes in the soil-vegetation system (Trabalka 1985). The total release since 1880 is estimated at 120 x 10^{15} g C (Solomon et al. 1985).

Table 7.1 Estimated values and uncertainties of key parameters in the global carbon cycle (from Solomon et al. 1985)

Parameter	Value	Uncertainty range
Net carbon flux from land biosphere since 1800	120 x 10^{15} g C	90-180 x 10^{15} g C
Gross annual terrestrial plant CO_2 uptake	120 x 10^{15} g C	90-120 x 10^{15} g C
Net primary production	60 x 10^{15} g C	45-62 x 10^{15} g C
Annual tropical forest area conversion (1970-1980)	0.004	0.003-0.006
Annual net carbon flux from land conversion (1970-1980)	1.3 x 10^{15} g C	0.0-2.6 x 10^{15} g C
Contemporary soil carbon	1.5 x 10^{18} g C	1.2-1.8 x 10^{18} g C
Contemporary biomass	560 x 10^{15} g C	420-660 x 10^{15} g C

During the last century the major component of this was probably the loss of soil organic carbon from the cultivation of virgin grasslands, but now it is from the felling of forests, mainly in the tropics. A major complication occurs in peatlands, where the storage of C can be regarded as part of a soil reservoir, or as a separate compartment (subfossil C). Indeed, the whole question of carbon changes in wetland ecosystems is extremely interesting (Olson et al. 1985), because of the potential changes in coastal wetlands due to sea-level rise, as well as to temperature change and the production of CH_4, another greenhouse gas, as well as CO_2 from such systems.

With an annual net flux from soil and plants of some 1 x 10^{15} g C and a total soil plus plant biomass carbon pool of around 2 x 10^{18} g C, there is plenty of scope for continuation or even acceleration of the present release of CO_2 from the terrestrial biosphere. The prediction of such changes is a matter of urgency.

THE ANALOGY APPROACH

There appear to be two ways of approaching the problem of predicting the impact of climate change on soils, which we define as the 'analogy' and the 'process' methods.

The analogy approach depends upon identifying a soil elsewhere which has developed under a climate that is very close to that expected at the site considered. There are several assumptions involved; the first is that climate is the dominant

factor in determining both vegetation and soil type. This may not be true if there are marked differences in parent material. Secondly, it assumes a fairly rapid re-establishment of equilibrium between soil and climate after a perturbation of the latter. In fact, different soil characteristics will alter at different rates (Fig 7.2), and the full development of a zonal soil in equilibrium with a climate takes thousands of years. Thirdly, this approach tends to assume that there is a relatively sharp change from one climate to another. In fact it is likely that this will be progressive, so that climate changes gradually, with soil and vegetation changes lagging behind to varying degrees. Nevertheless, with all these weaknesses, it is clearly a step forward to identify a soil that is the 'equilibrium' soil for a given climate, and then estimate the rate of progression towards this.

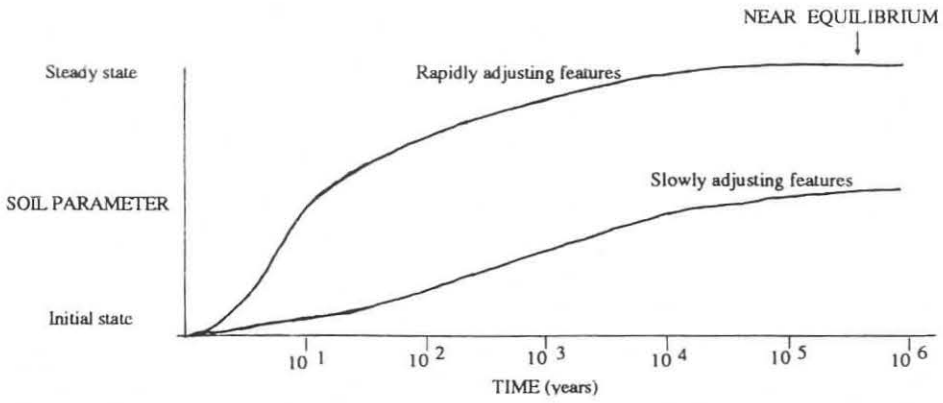

Fig 7.2 *Generalized diagram of the time taken to achieve steady state in soil properties after the initiation of soil development (from Walker and Graetz 1989)*

The vegetation classification of Holdridge (1947 1964) has been used to predict the alteration in vegetation zones that would follow climate change (Bolin et al. 1979). This analysis is based solely upon a projected temperature change scenario, and any additional changes, such as in rainfall, would certainly cause larger effects. However, given these vegetation changes, their impact upon soils can be estimated at least qualitatively.

One of the most interesting extensions of this system is the one of Zinke et al. (1984), who plotted the carbon content in 3400 soil profiles on the Holdridge triangular diagram (Fig. 7.3), to produce a set of contours. In principle, at least, the equilibrium amounts of organic matter in a soil with a defined climate can be determined, though it is hard to believe that other factors, such as the texture of the soil, are not involved. The predicted temperature changes will produce shifts on this diagram that imply a smaller total pool of soil carbon; it would be possible

to sum this change over the terrestrial surface. However, the structure of the diagram shows clearly that the soil carbon store is much more sensitive to changes in rainfall than to changes in temperature.

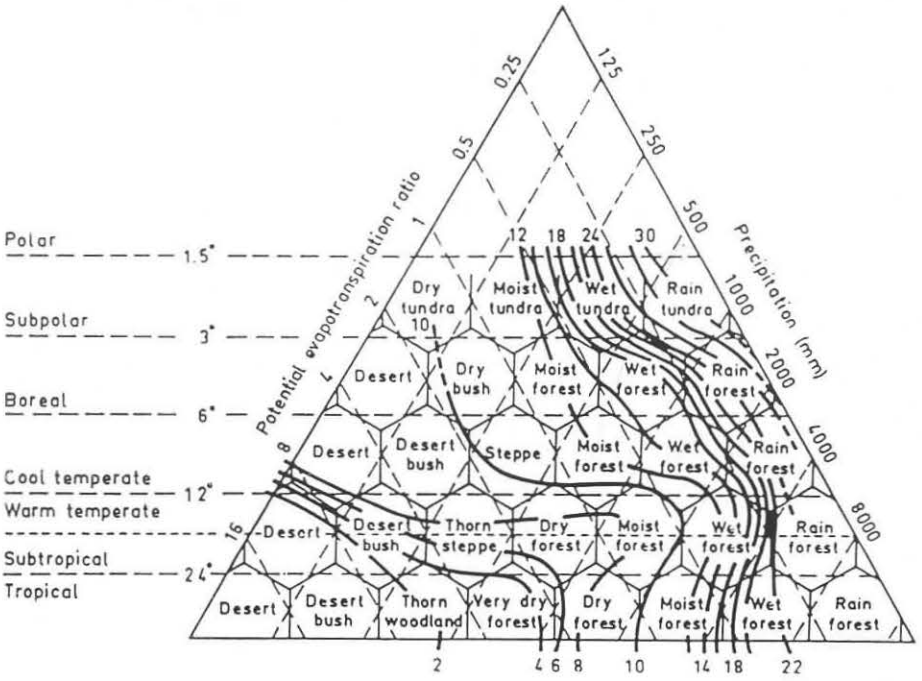

Fig. 7.3 Soil carbon content (kg m^{-2}) plotted as contours on a Holdridge (1967) life-zone chart (source Zinke et al. 1984)

It should be remembered that very large areas contain soils that cannot be appropriately delimited simply by the biome in which they are contained. In particular, cultivation, management and cropping change both the profile and the carbon content in all farmed soils.

THE SOIL ORGANIC MATERIAL BALANCE - THE PROCESS APPROACH

The components of the carbon cycle that determine soil organic matter levels (including subfossil remains such as peat) are:

- Net primary productivity of organic matter (NPP);
- Carbon partition above and below ground;
- Breakdown rate of organic debris;

- Formation of humus;
- Breakdown rate of humus.

The soil organic matter level in any soil will settle at an equilibrium level characteristic of its permanent characteristics, the climate and the land use or vegetation cover. This concept is most clearly shown in the studies of the soil organic matter levels in the long-term experiments at Rothamsted (Johnston 1987), in which any change in treatment caused a corresponding change in the progression to a new equilibrium level.

We draw a clear distinction between the decomposition of litter or debris, in which plant or animal structures can still be recognized, and soil organic matter, or humus. The latter is ultimately bound up with the mineral soil, has no recognizable structure, and has a fairly constant C/N ratio.

NET PRIMARY PRODUCTIVITY - INPUTS

In general, increased atmospheric CO_2 is expected to increase net productivity by more rapid photosynthesis. The present CO_2 concentration in the atmosphere, of ca 350 ppm, is suboptimal for plant growth in certain circumstances (Goudriaan and Ajtay 1979). The benefit to plants with the C_4 photosynthetic system is very small or zero, because of their lack of photorespiration. In C_3-plants there is undoubtedly the potential for increased growth, of perhaps 30%, if the CO_2 concentration rises to 600 ppm.

For a canopy or stand smaller values may be found than for single plants, because selfshading is less in the latter. In practice, growth of most crops or vegetation stands are limited by factors other than CO_2, e.g. drought, nutrient deficiency or disease. Even when such constraints are not overtly present, only a small fraction of even high yielding crops appear to reach the optimum yield set by radiation income (Tinker 1984). Any increase in crop yields is therefore uncertain, and major increases in net primary productivity in natural vegetation seem unlikely. However, if the growth constraint is due to water shortage, there may be a benefit from the more efficient use of water in plants whose stomatal behavior is controlled by CO_2 concentration. The increased CO_2 concentration in the atmosphere should then increase the ratio of the flux of CO_2 into the leaf to the flux of water out of the leaf, unless the relatively lower use of water raises the leaf temperature sufficiently to counteract most of the potential effect.

For a doubling of the CO_2 concentration, the water use per unit of photosynthesis could possible be halved (Squire and Unsworth 1988). Where water is limiting, this interaction of CO_2 level and water supply is likely to be the most important effect in practice.

There are other effects of increased atmospheric CO_2 which may be important in some circumstances. The morphology of plants may be changed, in

particular the root-shoot ratio. There is also evidence that the quality and composition of vegetation, such as the C/N ratio (Van Cleve et al. 1983), and also root exudation and mycorrhiza formation (Norby et al. 1987) is altered.

The problem in making any forecast concerning the climatic variables is the unpredictability of the latter. In general, temperatures will increase, and this will normally give a higher productivity, but the reverse is true if rainfall is not adequate. It is therefore impossible to predict net primary productivity at a designated spot in response to climate change, except to conclude that on average dry matter formation is likely to increase.

However, two general situations can be distinguished. Firstly, in most areas the general vegetation type will remain constant. There is a strong relationship between NPP and net annual evapotranspiration; the latter measures temperature, but with an inbuilt correction for any deficiency of water (Lieth 1975; Whittaker 1975) (see Fig 7.4). This can be used as a guide to the behavior of NPP in areas where the general vegetation type remains constant. Moderate changes in NPP in such conditions are unlikely to alter soil organic matter greatly.

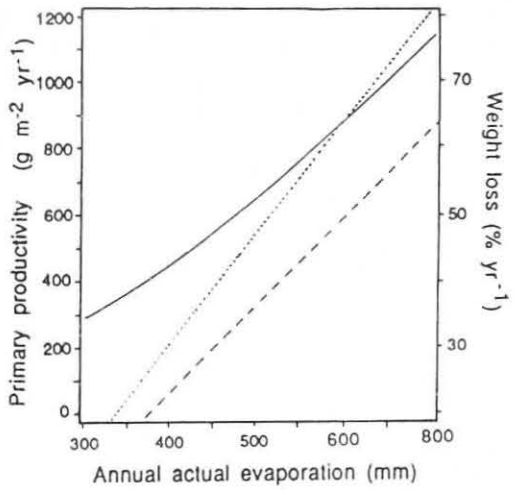

Fig. 7.4 *Relationship between actual evapotranspiration and productivity (——), decomposition of deciduous (– – –) and coniferous litter (·····)*

Secondly, the most striking effects on soils and vegetation will certainly occur near the boundaries of the present vegetation zones, where the whole character of the vegetation, and the type and amount of litter, may change (Table 7.2). Thus, Emanuel et al. (1985) calculated that the temperature rise from a doubling of CO_2 could result in a decrease in the global area of boreal forest and tundra of 37% and 32% respectively, this area changing into grassland or cultivated land. Soils in such conditions will then have quite new and different equilibrium states, with very different organic matter levels. The organic cycle

differs significantly in grassland and forest soils, the organic matter in the former being both greater in quantity relative to the plant weight, and more deeply distributed than in forest soils. The quality of the litter, and the type of soil fauna, also differ considerably (Anderson, in press). The 'analogy' approach must be used in predicting these.

Table 7.2 Plant biomass, litter input, soil organic matter and microbial biomass of major global vegetation types.

Subject	Tropical forest	Temperate forest	Boreal forest	Savannah	Temperate grass land	Tundra
Area x 10^{12} m^2	24.5	12.5	12.0	15	9	8
Plant biomass g C m^{-2}	18,000	14,000	9,000	1,800	1,440	250
Litter input g C m^{-2}	710	368	250	360	667	75
Soil C g m^{-2}	13,000	9,000	15,000	5,400	23,000	22,000
Soil N g m^{-2}	816	640	1,100	333	2,100	1,125
Microbial biomass C g m^{-2}	50	110	35	60	215	20
Microbial biomass N g m^{-2}	2	14	2.5	8.7	51	1
Microbial turnover yrs	0.07	0.30	0.14	0.17	0.32	0.27

Plant biomass and net primary productivity from Whittaker and Likens (1973). Litter from Ajtay et al. (1979). Soil C and N from Zinke et al. (1984). Microbal biomass from E.A. Paul (pers. comm.)

THE BREAKDOWN OF SOIL ORGANIC MATTER

The Q_{10} coefficient for the response of microbes to temperature change is normally around 2, but may vary from 1.6 to 3.2 (Schlesinger 1977; Singh and Gupta 1979). However, such figures are not necessarily a good indication of what happens when soil temperature is permanently increased in the field.

Microbial species, moisture content and resource quality will change, and all affect microbial reproduction and respiration rates. A laboratory measurement of the effects of a simple short-term temperature shift in a soil is therefore of rather little use, and only gives a very general indication of direction and magnitude of change.

The rate of decomposition of organic materials is sensitive to both temperature and moisture, and can therefore be related to the actual evapotranspiration rate in that environment. Hence, both this and the effect on NPP can be compared as in Fig. 7.4. Generally the effect of evapotranspiration on decomposition is greater than on NPP, hence as evapotranspiration increases,

there is a tendency for the ratio of vegetation biomass to litter to increase. This is very clear in a study of biomass and soil carbon in various biomes (Fig. 7.5).

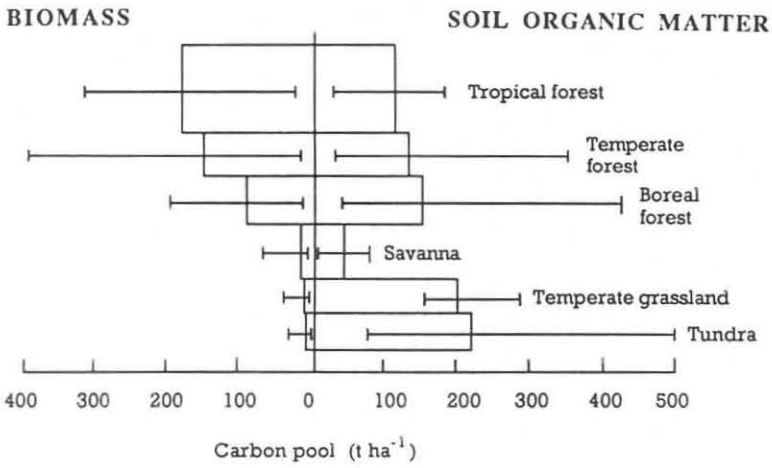

Fig. 7.5 Comparison of carbon pools in plant biomass and soil organic matter (including litter) across different ecosystems (from Anderson, in press)

The breakdown of organic material and humus has been successfully modeled. The clearest situation is in agricultural soils, in which the same crop is grown regularly year after year, so that input, in quality and quantity, remains constant. Jenkinson et al. (1987) have modelled this situation for the Rothamsted long-term experiments. The structure of the model is illustrated in Fig. 7.6.

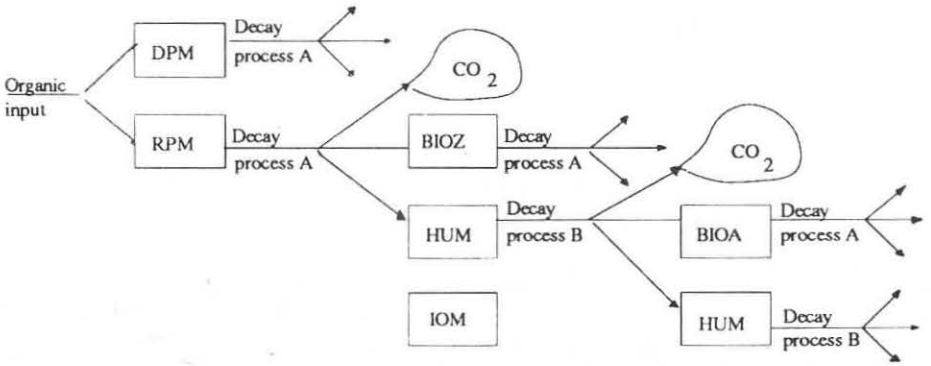

Fig. 7.6 Diagram showing the flow of carbon through the soil model of Jenkinson et al. (1987). DPM = decomposable plant material; RPM = resistant plant material; B10Z and B10A = zymogenous and autochthonous soil microbial biomass, respectively; HUM = humified organic matter

The ratio of biomass to humus produced is always the same in all soils, but differs between A and B pathways. The decay rate is $(1-e^{-k})$, where k is the rate constant, that varies with temperature and moisture (Van Veen and Paul 1981). The effect of soil texture is included by amendments of the ratio of CO_2 to biomass plus humus produced at each decomposition step. It was also found necessary to postulate a separate class of 'biologically inert' organic matter. This model has been applied to Nigeria Alfisols and Ultisols (Jenkinson and Ayanaba 1977), where it has successfully modelled the breakdown of 14-C labelled maize and rye grass residues, using the original parameters derived in Rothamsted.

A set of runs of this model are illustrated in Fig. 7.7 (Jenkinson, personal communication), with a direct comparison of the effect of two different temperatures, of 9.22°C (normal mean at Rothamsted) and 19.22°C. The higher temperature leads to more rapid decomposition of residues, as expected. Broadly, the amount remaining at any time at 19.2°C is roughly equivalent to what is left at the lower temperature, but after about 3 times the time.

Using the effects of change of biome, of changes in CO_2 and of actual evapotranspiration on NPP, and the existing models of organic decomposition, it should be possible to make prediction on the soil carbon levels following climate change. The uncertainties and errors will however be very large.

SOIL ORGANISMS

The general soil heterotrophic population will behave in a way largely characterized by the speed of decomposition of soil organic matter. Temperature changes and precipitation changes will have effects as discussed above. Species of micro-organisms will change, but the likely speed of change, and the acceptable climatic limits for any one group pf organisms, are such that no major unexpected effects are likely. The temperature optima for soil organisms are usually quite broad, and a shift of one or two degrees will normally make little difference. The most interesting problem in this topic will be in the wetlands of the high latitudes of the northern hemisphere. Temperature changes will be largest here, perhaps 6-10°C, with sharp changes in water relations. The question of whether the microflora will change in line with the conditions, or will lag well behind these, does not appear to have been addressed yet. The long-distance transport of bacteria on migrating birds, animals or man will normally be fairly rapid; this is less certain with the larger soil organisms.

Previous comparisons of microbial communities across different biomes have been largely mycological, and show that species composition is clearly related to biome type, even for the same biome in different continents (Kjoller and Struwe 1982). Simple biomass comparisons by these authors suggest that upper levels for fungal biomass are an order of magnitude higher in temperate grasslands than in tundra, and that turnover rates are similarly higher in the more

productive systems. Bacterial populations have been less well studied in this respect, yet Sundman (1970) has demonstrated that the functional attributes of the soil bacterial population are influenced by plant cover type.

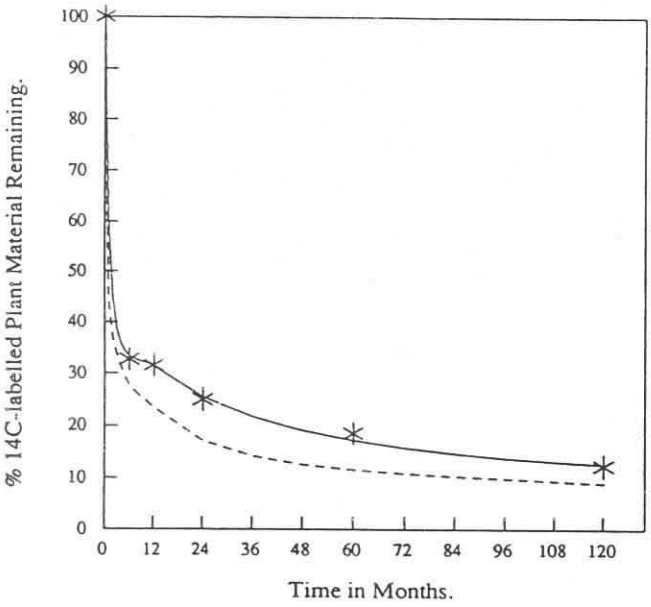

Fig 7.7 Runs of the Jenkinson et al. (1987) model, showing a direct comparison of the predicted rates of decomposition at 9.22 °C(——) and 19.22 °C(- - - -)

There is no direct evidence relating faunal biomass or productivity to organic matter quality; these seem to reflect responses to the environment directly. Thus, for example the distribution of termites appears to be limited by low minimum temperatures (Harris 1971), and the increases in temperature predicted as a consequence of climatic change can be expected to extend their range.

Soil organisms with more specific characteristics may well be slower in adjusting to the new conditions. These are:

- Plant pathogens;
- Symbiotic organisms;
- Soil fauna.

For plant pathogens and symbiotic organisms there is the additional complication that they are, to varying degrees, dependent upon specific types of vegetation. A vegetation zone shift caused by temperature/precipitation changes will only have reached full equilibrium when both the vegetation and the appropriate lower organisms have established themselves together.

In terms of temperature only, the North-South displacement equivalent to the predicted climate change is up to a few hundred kms, depending upon latitude. With a CO_2 doubling time of 40-50 years, the rate of movement of organisms may need to be up to 10 km yr -1 to keep up with this shift. There is some information on the rates of spread of soil borne diseases and the recent rapid movement of the barley yellow mosaic virus across England provides an example of the time scales involved (Hill and Walpole 1989). In the seven years from the initial reports of the occurrence of this soil borne virus in England, the organism is now established over much of the country, and has spread at a rate which clearly exceeds those anticipated for climatic change. However, there are no precise analogies with the proposed scenario, in that all studies of organism dispersion have been within a broadly static environment, following an introduction of the organism. There has been no real analogy since the end of the last Ice Age, when the rate of spread and change was probably (though not certainly) slower.

It will be essential for the appropriate symbionts to be available to the invading vegetation. Amongst the mycorrhizas, there is likely to be a decrease in the amounts of ectomycorrhizal fungi, as the large boreal forest belts move northwards towards the poles. Tropical and subtropical trees are almost wholly symbiotic with vesicular-arbuscular mycorrhizal fungi (Harley and Smith 1983). The rate of movement of mycorrhizal infection from root to root in grassland is slow, at ca 50 cm/yr (Sparling 1976). Fortunately these have low specificity, and some common species are found very widely distributed. Except in very high latitudes or in previously bare soils acceptable mycorrhizal symbionts are likely to be present, though the best equilibrium population of fungi may take a considerable period to establish. The N-fixing Rhizobia, actinomycetes or free-living bacteria are more specific in their hosts, in particular the first two. There will certainly be situations in which there is a delay in introduction of these symbionts, and the spread of some leguminous plants may be delayed because of that.

Heal and Ineson (1984) provide evidence to support the argument that the pattern of primary productivity in different biomes is reflected in the microflora and fauna, mainly through variation in resource quality, but also particularly for the fauna, through external climatic factors for the fauna. Again, it is when climatic change results in vegetation changes that associated shifts in below ground populations will become most pronounced.

The time taken for the soil fauna to migrate into new areas is expected to be slower than for soil microorganism, and introductions of, for example, earthworms have shown rates of migration restricted to a few meters per year, in the absence of appropriate vectors (Van Rhee 1969). The net result is that the rate of change in climate may exceed the rate of colonization by the soil organisms usually associated with the new climate, and this lag will be greater for the larger components of the soil biota. The longer term picture is shown in Fig.7.2,

suggesting that equilibrium conditions will ultimately be reached; it is the predicted speed of change of climate which will result in imbalance.

LAND CAPABILITY EFFECTS

Over most of the tropical and subtropical parts of the world, the restraint on plant production is water or fertility rather than temperature or radiation. As temperature will change relatively little with climatic change in these latitudes, the overwhelmingly important effect is likely to be in rainfall changes, which are the most difficult to predict. Any rise in temperature would be expected to produce more intense rain, even if the mean precipitation were unchanged or were lowered, with consequential effects on infiltration rates and erosion.

In the higher latitudes in particular, there may be major changes in land use, with the movement of crop belts to the North (Warrick et al. 1986). If this involves the opening up of previously uncropped grassland or forest, the organic carbon contents of the soils will decrease, which has already contributed to global CO_2 increase (Trabalka 1985).

The greatest potential for catastrophe must be in impacts on the dryland areas of the world. These are soils which are already infertile, often with serious additional constraints of poor drainage, wind or water erosion, stoniness or shallowness and salinity (Dregne 1982). Their agriculture is already often barely viable, and dryland areas contain some of the most disastrous agroecological problems in the world (Steiner et al. 1988). Any decrease in mean precipitation, or any increase in rainfall irregularity, could be very damaging in many areas. Even now, the rainfall distribution is normally skewed, with many more years below the mean than above. The risks associated with this type of farming can easily become so large that the whole system loses its viability.

The degradation that already occurs in many of these soils, will be accelerated by and associated with any decrease in organic matter (Lal 1987). The lowering of dry matter yields by lower rainfall will cause lower residue returns to the soil. A temperature increase will lead to some acceleration in organic matter decomposition, and therefore soil fertility (most probably nitrogen supply), water retention, infiltration and erosion resistance will gradually diminish. Changes of this type may be catastrophic in such areas, and the effects on soil organic matter will be strongly implicated.

CONCLUSIONS

1) Climatic changes may have massive consequences for the total vegetation/soil system , and it is quite unprofitable to consider soils alone in this context. The impacts will vary widely with the actual climatic change

experienced, and also with the sensitivity of the vegetation-soil system at any point.

2) The changes are likely to result in a significant net release of carbon dioxide to the atmosphere, and it is important to be able to predict this as soon as possible. Hydrological and vegetation changes may be more important than simple temperature changes in this regard.

3) Scientists dealing with the terrestrial part of the global system need to develop a more integrated approach, such as is particularly evident in the atmospheric sciences. Single site studies may be useful in defining principles, but are of little value on their own in developing a global picture. Soil scientists in particular need to emphasize the global nature of soil science, not so much in taxonomic pedology, but in process studies. There is a fundamental need to improve our methods of extending data from point measurements to global fluxes and processes. We need to look very carefully at our use of current classification systems, that are mostly based on semipermanent, slowly changing characteristics. We also need to improve our use of remote sensing technology. Where new technology or instrumentation are relevant, we should have no hesitation in identifying this as an urgent need.

4) This short review has shown quite clearly the weaknesses in soil ecology, and that additional effort and resources are needed if better predictions are to be made. The comment applies particularly to microbial ecology, in which the vast majority of soil microbes cannot even be identified, much less studied. It also applies to the ecology of soil fauna.

5) More work should be done on soils that are at the boundaries of vegetation zones, and are thus sensitive to change; in general, soils not in equilibrium with their environment warrant more interest.

6) Better methods of studying organic decomposition are needed including standard materials for testing rates (Anderson, in press).

7) We must recognize the weaknesses of GCM predictions, but be ready to seize upon every advance as it appears.

ACKNOWLEDGEMENTS

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